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WHAT GRAVITY TELLS US

Kurt Lambeck and Francois Barlier*

ABSTRACT. Certainly the earth is not round. It is an ellipsoid but its gravity field does not vary according to its shape and its inevitable irregular features. From this, how can it become known and what conclusions can be drawn on the detailed structure of our planet?

The science traditionally associated with the problem of determining the terrestrial gravity field is called geodesy. Measurements required for these studies have been made over the surface of the earth, but the utilization of artificial satellites has resulted in considerable improvements. It may seem paradoxical that we have to remove ourselves from the earth in order to obtain a clearer picture of our planet, of its shape and interior. Satellites allow us to obtain a global view whereas classical observations made on the surface give us a more regional picture. /24**

In general, it is assumed that the determination of the gravity field is based on static concepts. However, measurements have become so accurate that, in order to determine the quantities related to these problems, it is necessary to take their variations with time into account. Earthquakes on the earth prove that the platform on which we are carrying out our measurements is

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unstable. The finno-Scandanavian rise and the relative movement of enormous pieces of the earth's crust, which can amount to 10 or 15 cm per year, represent other examples. In the same way, the potential field of the earth cannot be considered as constant. The earth is not perfectly rigid. It is continuously deforming under the influence of the attractions of the moon and the sun. There is a change in the mass distribution, and consequently the potential field varies as a function of the earth-moon-sun geometry.

These examples show the necessity of considering the earth in a dynamic way when the problems mentioned above are studied. Consequently, if we follow a methodical observation program, we can hope to observe all the effects of the internal and external forces on the surface and to finally obtain a picture of what goes on in the interior of the earth (Figure 1). It should be understood that we must relate these results to other types of information, such as the heat flux, seismology, magnetism, tectonics, as well as experimental and theoretical geophysics.

In general, in a discussion of this type, the methods are first explained, then the results and the geophysical implications. We have decided not to follow this order, and instead we will commence our discussion by asking ourselves what our knowledge of the earth's gravitational field can tell us about the interior of the earth. We will also discuss the information necessary to verify certain hypotheses or theories. Finally, we studied the methods and the improvement to the methods necessary to determine a better model of the earth.

ANOMALIES OF THE REAL FIELD

In the first approximation, we consider the earth as an ellipsoid. The greatest irregular features in the average level of the oceans with respect to this ellipsoid are about 100 meters. This first fact is already of great importance for geophysics. In effect, the ellipsoid of revolution is the surface that a homogeneous mass of fluid takes on when it rotates. Thus, we have /25
good reason for believing that, given the shape of the earth, the high

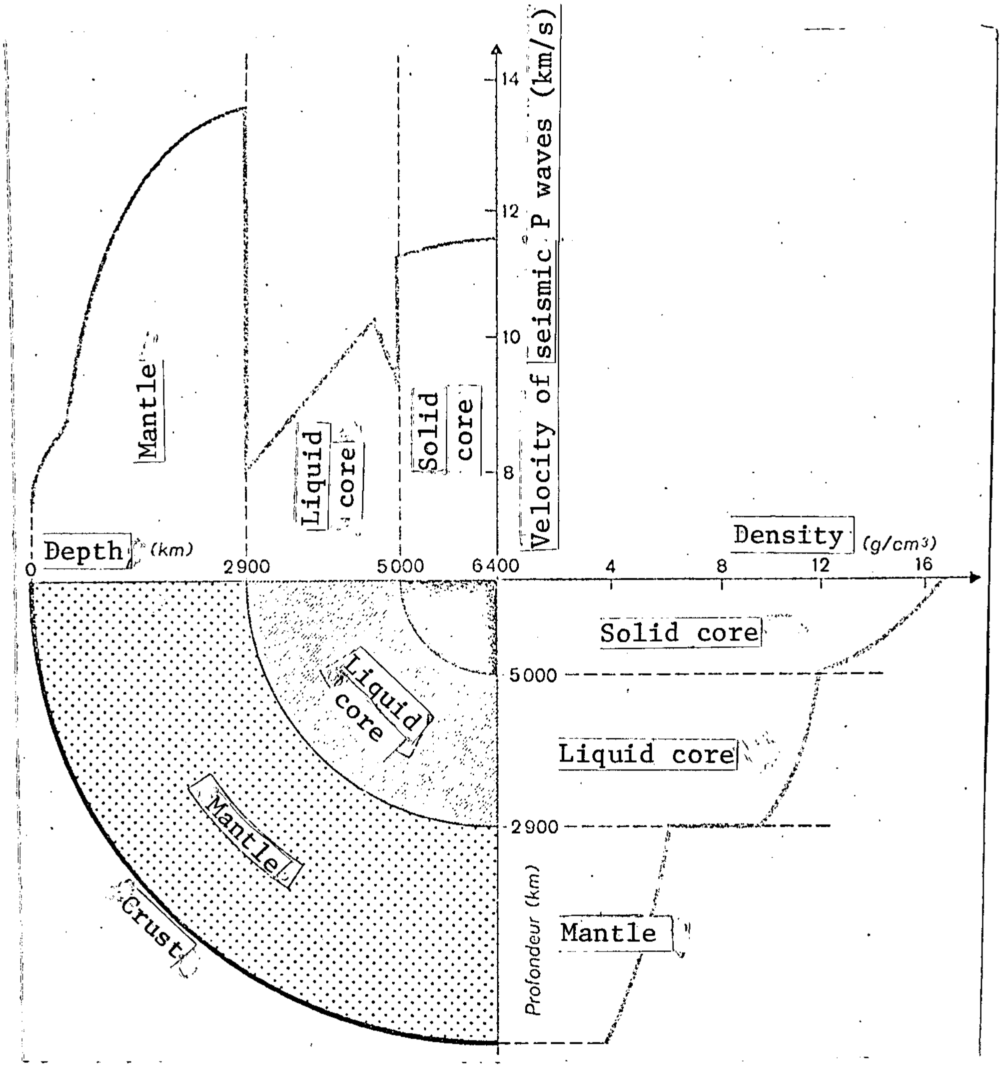


Fig. 1. By studying the various waves emitted by various seismic events by means of observation stations distributed over the entire earth, it has been found that the first waves emitted, the P waves, depend on the depth at which the waves are moving. In particular, it has been noted that this velocity varies considerably and in a discontinuous manner at depths around 2900 and 5000 km. These observations, together with seismic observations, have made it possible to establish an approximate model of the earth's structure. Below the terrestrial crust, two zones can be distinguished: One region called the mantle which extends down to 2900 km depth; the rest of the earth makes up the core. The properties of the latter change around a depth of 5000 km. The seismic observations have also made it possible to calculate the average density of various regions (horizontal axis) as a function of depth (vertical axis). As we move deeper into the earth, this density increases gradually toward the interior within each of these three zones. This increase is very large when passing from one zone to the other, especially between the mantle and the core.

stresses within it will make it behave like a body of fluid. Under these conditions, we can theoretically study the problem of a fluid of homogeneous density which is in uniform rotation, where the characteristics, density, rotation parameters, flattening are those of the earth. The flattening of such an ellipsoid, that is, the ratio of the difference between the diameters at the equator and at the pole to the mean diameter, must be 1/231. The ratio derived from satellite trajectories is approximately 1/298.25. From this, we must conclude that the earth is not homogeneous. Nevertheless, the law which defines the density as a function of depth cannot be an arbitrary one. In the first approximation, it must satisfy the equilibrium equation of a liquid column under the action of its own weight, hydrostatic equilibrium. Astronomy gives us other constraining equations regarding the mass distribution, i.e. regarding the moments of inertia of the earth. These come above from the precession of the equinoxes. The latter is due to the torque which the moon and the sun exert on the earth. This couple would be zero if the earth were a sphere. Under these conditions, one finds that the flattening of the earth would have to be approximately 1/300.

In reality, the gravity field is more complex and we can represent it by means of deviations from a model, i.e. by the differences between the real field and the field of a reference ellipsoid. These deviations are called gravity anomalies (see Figures 2 and 3).

If the earth is not in hydrostatic equilibrium, two interpretations are possible for taking the anomalies into account. It can be assumed that the earth is sufficiently rigid to support pressures and stresses brought about by the anomalies in the gravity field without carrying out a development for a time interval on the order of several dozen million years. On the other hand, it can be assumed that the earth is not very rigid, but the density anomalies can be maintained by thermal convection within the interior of the earth.

It appears that the following two plausible interpretations coexist with each other: depending on the part of the earth which must be considered responsible for the anomalies, one or the other of them will prevail. Seismology

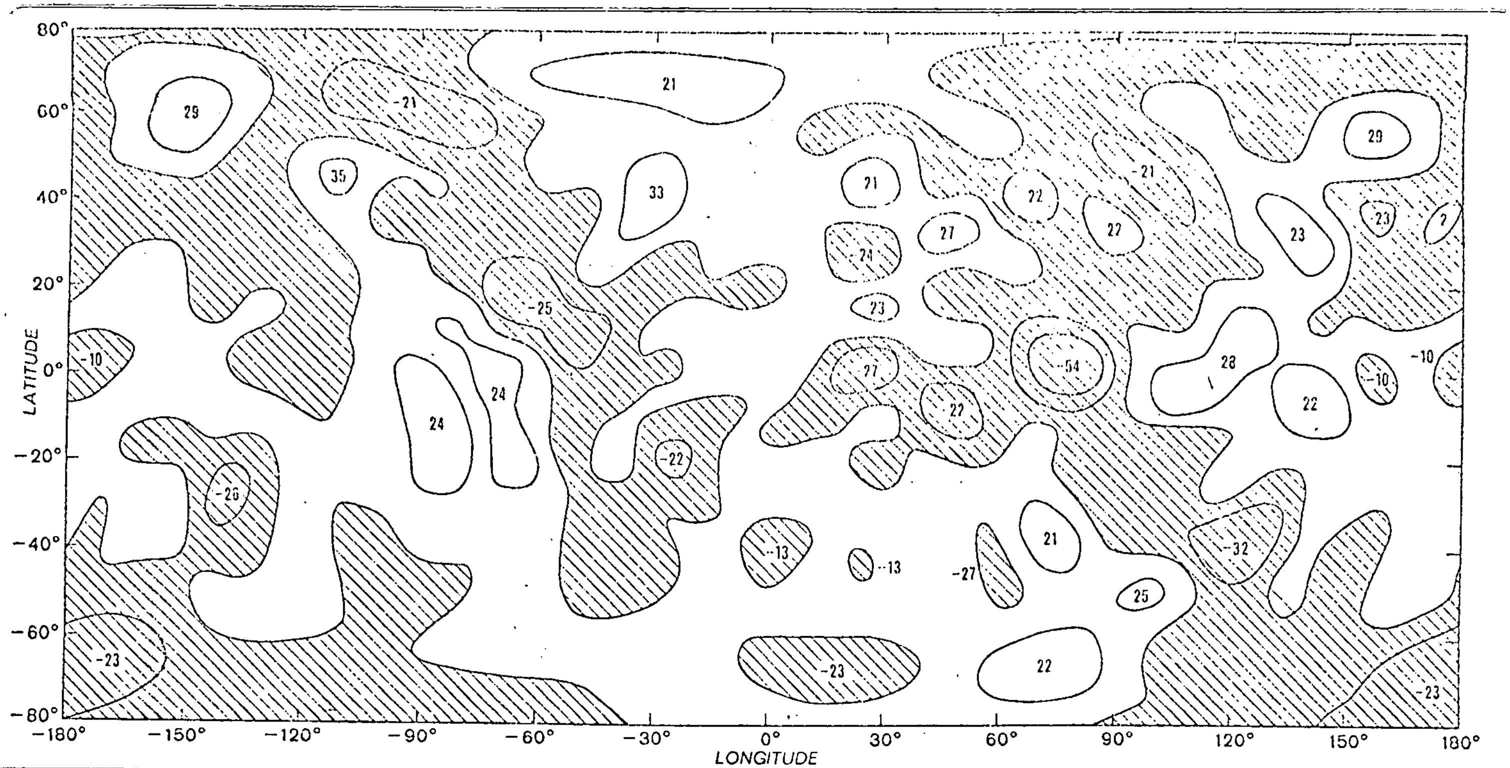


Fig. 2. In the first approximation, the earth is considered as an ellipsoid. The flattening which is measured, that is, the ratio between the difference in the diameters at the equator and the pole, to the mean diamter, depends on the method used to make these measurements. Because of analysis of artificial satellite trajectories, geophysicists Gaposchkin and Lambeck found a value of 1/298.25 for this flattening. When the gravity field is measured at each point of the earth's surface, one does not always find exactly the same value as would be found by calculating with an ideal ellipsoid. The difference between the measured value and the theoretical value, called reel field anomaly, depends on the location where the measurement is made; it can be positive, negative or zero. The map above shows the distribution of gravity anomalies for an ellipsoid with a flattening of 1/298.25. Level curves are given in steps of 0.02 cm/sec^2. The shaded areas correspond to negative anomalies. Numbers in the center of each curve correspond to the highest value of the absolute anomaly value within the region bounded by this curve.

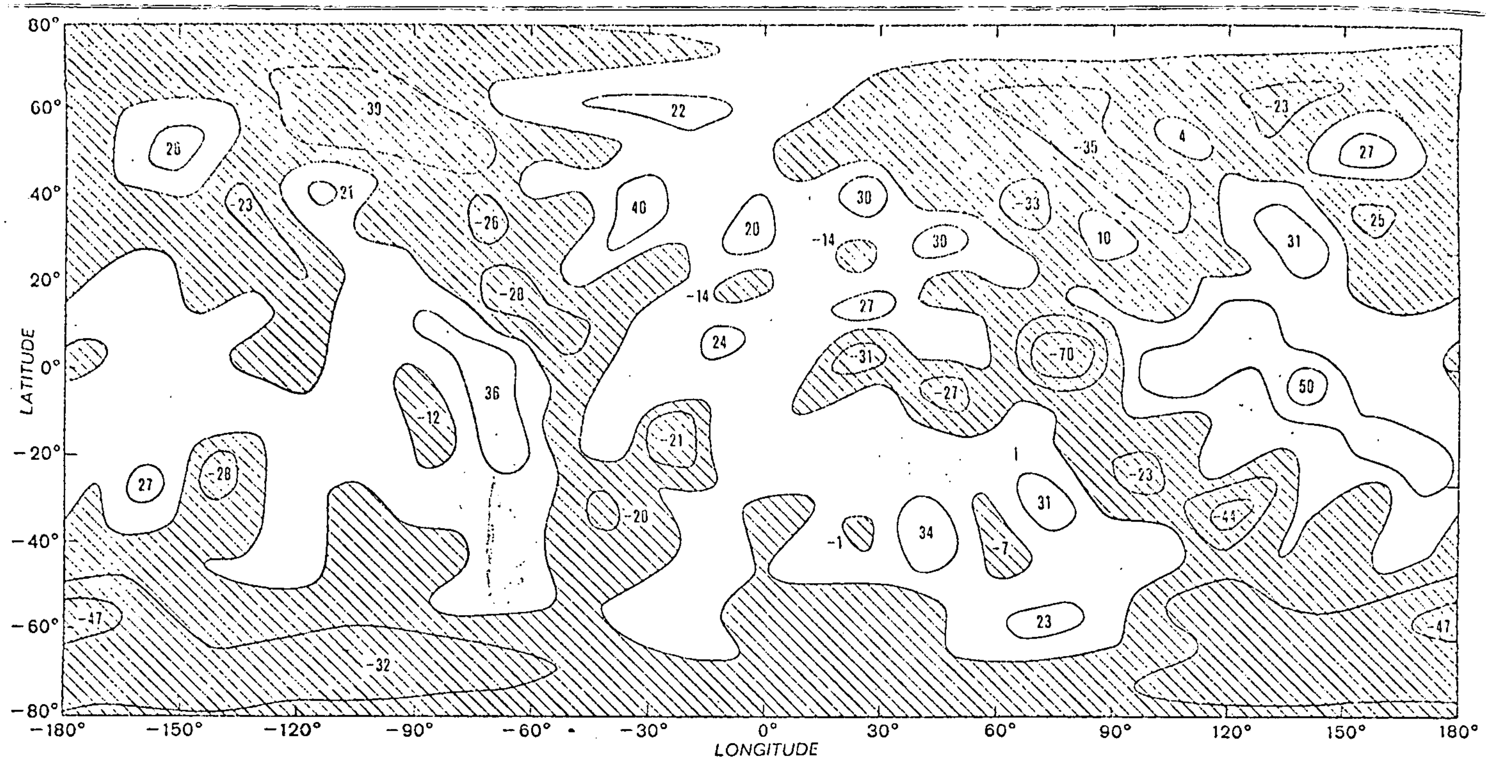


Fig. 3. If we consider the earth as an ellipsoid which is sufficiently liquid so that each of its parts is in equilibrium under the action of its own weight, i.e. hydrostatic equilibrium, it is found that the flattening of this ellipsoid is 1/300. The gravity anomalies measured for this ellipsoid are shown here and are somewhat different from those for the ellipsoid having the flattening 1/298.25 (see Figure 2). We can ask whether these differences are sufficiently significant to merit a special interpretation.

indicates that the core is sufficiently liquid so that it does not deviate in a significant way from hydrostatic equilibrium. For this reason we will focus our intention on the role which the crust and the mantle play in the observed anomalies.

FLATTENING AND VISCOSITY

The first characteristic of the gravity field is translated into a difference between the observed flattening, which is on the order of 1/298.25 and the flattening of an ideal earth model in hydrostatic equilibrium, for which the flattening should be 1/300. Certainly the difference is small, but it is /26 larger than the errors with which both quantities were determined. The observed flattening is too large by about 0.5%. The question is to find whether this difference is sufficiently significant with respect to the other anomalies (Figures 2 and 3) to merit a particular interpretation.

W.H. Munk and G.J.F. MacDonald established the hypothesis that the present flattening of the earth corresponds to the flattening which it had during geological times when it was rotating faster. If we assume that the mantle is elastic, incompressible and that it can resist certain deformations, we can calculate the stresses which must prevail within the interior of the mantle. One finds pressures on the order of 200 atmospheres which apparently are too high for the material which makes up the mantle. The mantle itself is at a high temperature.

If we also assume that the earth behaves like a viscous body and that its rotation is slowing down because of friction in the oceans, we can show that the flattening will progressively decrease, even though when there is a certain decrease, this flattening will readjust itself to the new rotation rate. The slowing down of the earth's rotation can be calculated from the increase in the duration of one day. The day increases by one or two seconds every 100,000 years beginning at least 300 million years ago. From this, one can derive an estimate for the retardation which is necessary so that the earth's flattening will readjust to the variation in the earth's rotation rate. This retardation will be on the order of 10 million years. Therefore, it is estimated that the viscosity of the earth's mantle is on the order of 10^{26} poises* which may seem high compared with anything we know on the surface of the earth. This value must be representative for the main part of the mantle. This probably precludes the existence of large areas in which convection is present and a large shifting of the axis of rotation.

The viscosity of the mantle is probably the most enigmatic characteristic of the earth, because the estimated values of it are always based on certain hypotheses. Another method consists of measuring areas which were covered with ice during geological times, such as Finland for example, or to measure lakes which have recently evaporated. A viscosity between 10^{21} and 10^{23} poises is found depending on the assumed depth at which the rising originated. These numbers only hold for the first 1,000 kilometers of the mantle. They differ by several orders of magnitude from the numbers given before.

Other geophysicists, such as P. Goldreich and A. Toomre, believe that the observed difference in the flattening is not as significant as the other deviations from hydrostatic equilibrium (Figure 3). According to them, a separate interpretation of the various anomalies could be misleading and unjustifiable. /27
The earth must be considered as a viscous body turning around an axis in such a way that its moment of inertia with respect to this axis is a maximum. If during the geological ages, the gravity anomalies had been distributed differently, which is suggested by the drift of continents, the axis of rotation would have to be located at a different position. With this interpretation, it is not necessary to introduce the time constant of approximately 10 million years which was derived from the slowing down of the earth's rotation. Instead, a time constant is derived from the displacement of the axis. If the axis is at the present time displaced by 10 kilometers over millions of years, as paleomagnetic data show, we obtain a viscosity measurement of about 10^{24} poises.

*Poise: unit of dynamic viscosity. Water, for example, has a viscosity on the order of 10^{-2} poise.

This value is in better agreement with the value derived from the recent rising of various areas, and it reintroduces the possibility of thermal convection in the lower mantle.

However, the information obtained from the shift of the pole is based on a certain number of hypotheses, the validity of which can be contested. It is assumed that the lines of force of the magnetic field have not changed in the course of time. The second hypothesis is that the field was coupled with the rotation axis of the earth in geological times. These two hypotheses are not completely demonstrated by paleomagnetic data.

The incompatability of these two conclusions shows how dangerous it is to introduce hypotheses without sufficient information. The average gravity field is probably better known over the entire earth than other geophysical quantities, and the knowledge of the earth's flattening leads to two considerably different interpretations.

IN THE MANTLE: THE CAUSE OF THE ANOMALIES

If we accept the hypotheses of Munk and MacDonald regarding the flattening of the earth, we should analyze the other gravity anomalies with respect to the ellipsoid having a flattening of 1/298.25, because the other hypothesis requires an analysis with respect to the ellipsoid having a flattening of 1/300. In either case, it is remarkable (Figures 2 and 3) that there is no correlation between the gravity field and the distribution of the continents. This leads to the important conclusion that the density anomalies which the gravity anomalies produce are not related to the continents but have their origin farther inside the mantle. The fact that there is no correlation represents a practical demonstration of the isostasy hypothesis which says that the continents are floating on a denser material.

The density increases from 6 to 10 g/cm^3 (Figure 1) between the core and the mantle. This increase takes place over a transition zone which is only possible 5 kilometers thick. This is why we can direct our attention to establishing density anomalies within this layer which are the origin of the anomalies observed on the surface. An indication of the existence of such anomalies comes from an interpretation of the earth's magnetic field and the variations which are observed in the earth's rotation rate. Some simple calculations show that only global scale phenomena can be caused by anomalies within this layer. Thus, the residual gravity field after subtracting the anomalies assumed to have their origin in this transition zone (Figure 4) can be represented in this way. /29

On the other hand, seismology indicates that the mantle is reasonably homogeneous in the lateral direction from the core up to about 1,000 kilometers below the surface of the earth. No significant variation in the density has yet been found, and it is possible that large convection zones exist over this entire region. It therefore seems that this is not the region in which to look for an explanation of gravity anomalies (Figure 2). On the other hand, if we consider the upper regions of the mantle, it seems that the density variations could exist and that the post-glacial rises give us a sure indication of the possibly existence of very slow flows of matter.

It is known that the lithosphere, the external and rigid envelope of the earth, consists of plates bounded by ocean crests, ocean depressions and faults. These plates slide over the more fluid asthenosphere (see "From the Renewal of the Ocean Floors to the Revival of Geology", Science Progrès Découverte, No. 3437, October, 1971).

The motions of these plates can be estimated from the marine magnetic anomalies: one finds quite high values, such as 10 cm per year.

If we compare the distribution of the gravity anomalies (Figures 2, 3 and 4) with the tectonic features of the earth (Figure 6), one finds a high correlation between the strongly positive anomalies in the gravity field and the ocean depressions. However, in the Alpine belt, the correlation is positive in the western region but becomes negative in the eastern part. One finds a correlation with the ocean crests which is analogous, but in general it is

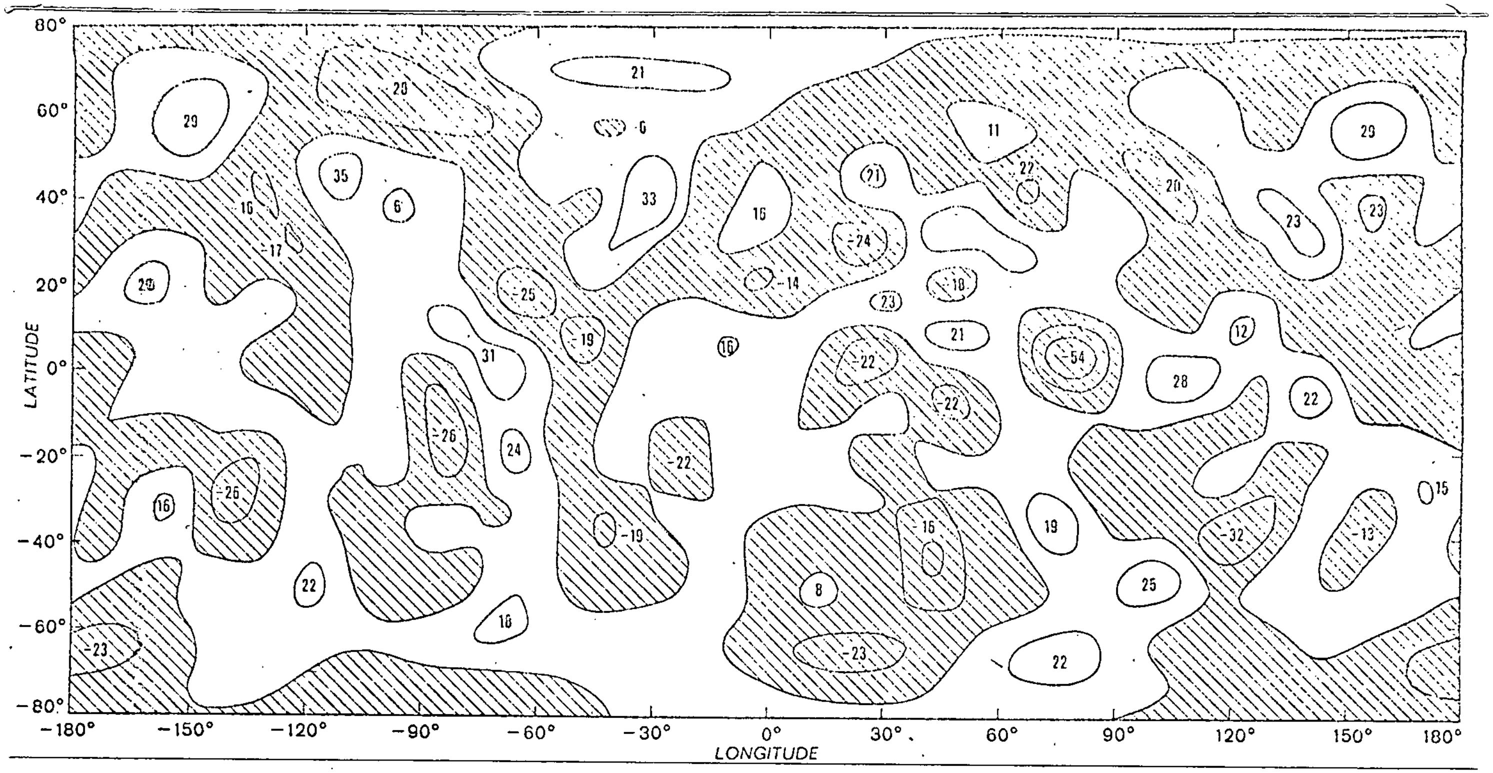


Fig. 4. Within the transition zone between the mantle and the core, at a depth of about 2900 km, the density varies wildly (see Figure 1): this zone could be the reason for certain gravity anomalies which can be calculated. By eliminating these anomalies from Figure 3, one obtains the map shown above. The residual anomalies shown here will be due to phenomena much closer to the surface.

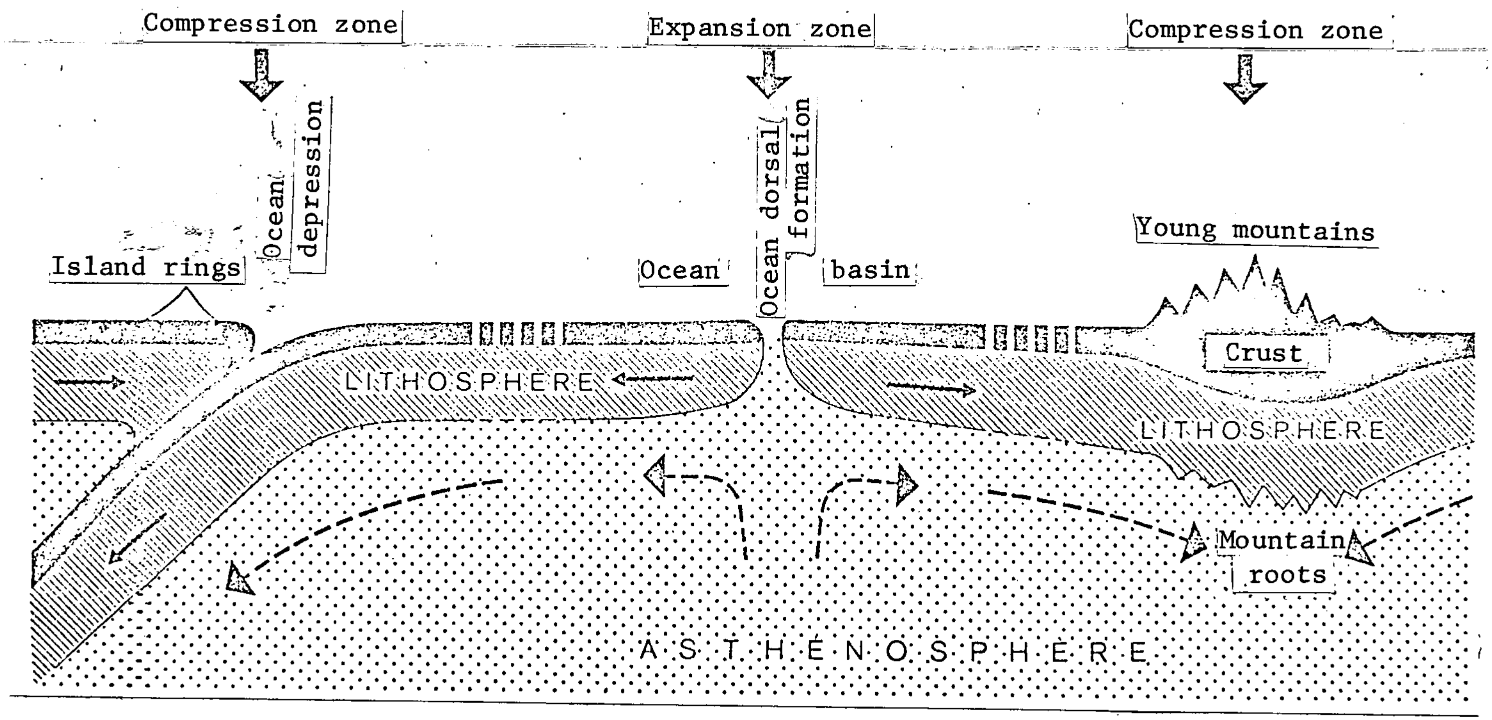


Fig. 5. There are compression and expansion zones at the boundaries of the plates which make up the lithosphere. In this schematic cross section of the earth's crust, the solid arrows correspond to movements of the lithosphere and the dashed arrows correspond to the movements of the astenosphere. The compression zones can correspond to either ocean depressions with surrounding island rings, or to mountain chains. The expansion zones are connected with the ocean dorsal formations. The ocean basins, regions within which thick sediment layers are found, are often located between the dorsal regions and the compression zones. The extent of these basins is very large compared with the thickness of the compression and expansion zones. The region occupied by these basins is indicated by a dashed line in order to show the scale change.

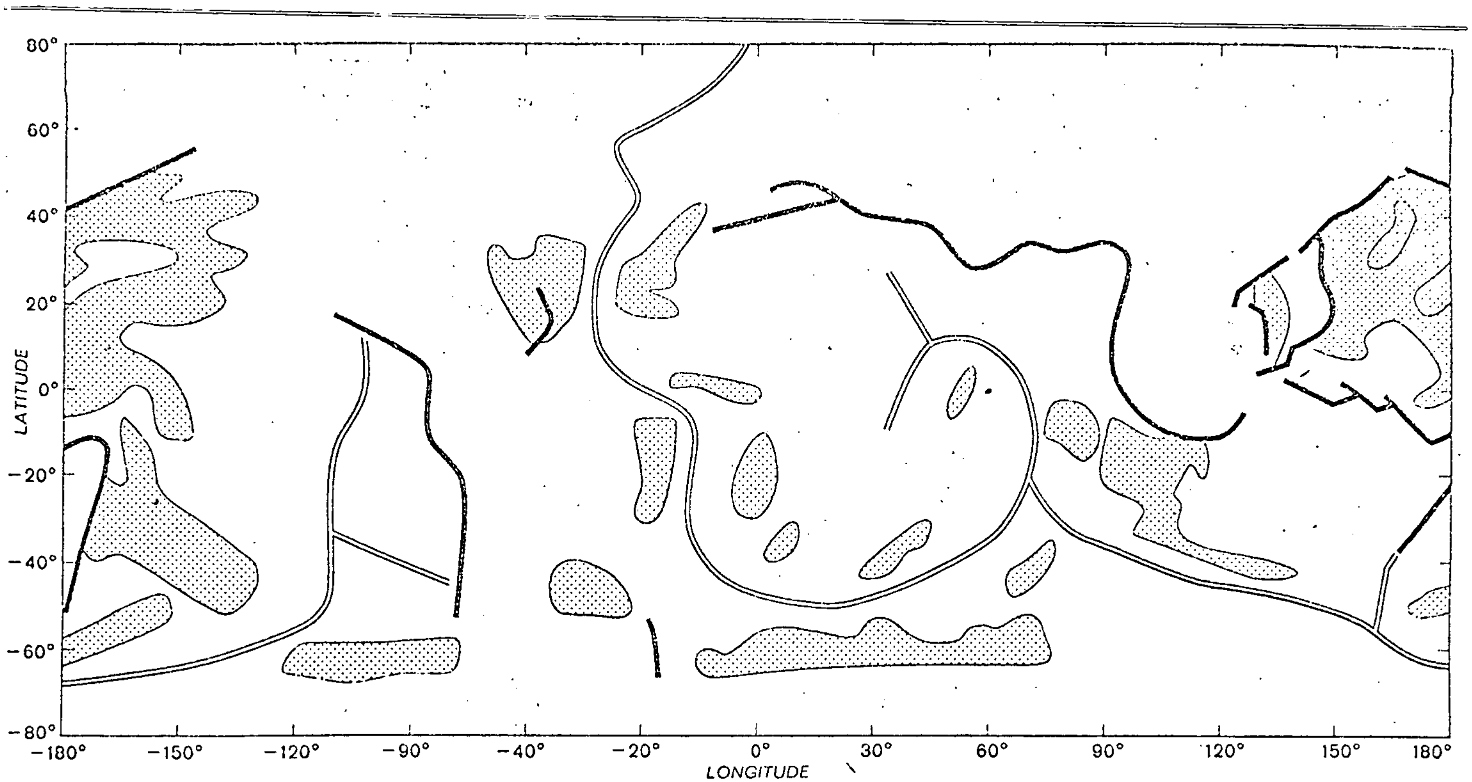


Fig. 6. A very schematic representation of the distribution over the earth of compression and expansion zones at the boundaries of the plates making up the lithosphere. These zones are limited on one side by compression lines (solid lines) which line up with the island rings, for example, in the region of Japan, or with mountain chains, such as in Eurasia. On the other hand, the expansion lines (shown as double lines) follow the ocean dorsal formations, such as is the case in the middle of the Atlantic. The dotted regions represent ocean basins.

smaller and sometimes even negative. The ocean basins are zones where gravity anomalies seem to be mostly negative.

The geophysicist W.H. Kaula attempted to interpret the correlations as being due to quite complex interactions between the convection flows in the asthenosphere and the lower boundary of the lithosphere. For example, the existence of positive anomalies below the depressions would be a consequence of the sinking of the colder ocean lithosphere plate below the continental shelf and of the reaction of the asthenosphere which pushes the lithosphere upwards. The positive anomalies which are found below the ridges can be explained if the material pushed from the interior to the exterior (which is warmer than the regions which it has left and, therefore, less dense) has sufficient mass to bring about an increase in the gravity. Recently, calculations regarding this point have been made using highly simplified models of the asthenosphere which did not take the lithosphere plates into account. In these models, one finds that the anomalies are positive in regions where material is pushed upwards. The presence of the lithosphere, which is a relatively thin layer above the dorsal formations, cannot greatly change the conclusions of this theoretical study, which can be generalized. /30

The mechanisms which we have described are only very approximate ones. It is necessary to carry out calculations which take into account the orders of magnitude of the existing forces in order to define accurate models and to understand the gravity anomalies and the systems of convection flows. These calculations are very difficult because we do not know the properties of the mantle and the lithosphere well, and we do not know where the lithosphere conceals and modifies the characteristics of the asthenosphere. This is why it is necessary to use other geophysical, seismology, heat flux and other data in order to study the problem described above in a valid way. Unfortunately, these data are still scanty, especially on a global scale.

HARMONICS

In order to show that the problem of determining the gravity field can be simplified by the use of series (which are infinite in theory) we can proceed as follows:

Let us consider the earth's equator. At each point we can measure the gravity. Assume that we want to represent the gravity field as a function of distance from an arbitrary origin point. We can write the gravity in the form of the following functions: a constant equal to the average value of gravity plus a sine function which will have two maxima along the equator, plus a sine function which will have four maxima and so on. In front of this function, there is a coefficient which we will call harmonic coefficient. The distance between consecutive maxima is called the wave length of the harmonic. It can be shown that the influence of a harmonic on the motion of a satellite is decreased not only when the altitude is increased, but also when the wave length is decreased. This results in simplifications in the calculation.

Let us now consider another aspect of the consequences of these coefficients regarding the motion of the satellite. If the gravity field were a very simple function, i.e. equal to a constant, the satellite would describe an ellipse like that of the earth around the sun. In fact, the field we are describing is more complicated and the motion is described as the sum of various functions: a principal function corresponding to the elliptic motion plus periodic functions the periods of which range from the time interval necessary for the satellite to perform one revolution to fractions of this period.

We also know that the amplitude of these functions with a small period is small. These amplitudes cannot be observed. In effect, using optical observations and laser distance measurements, we can only observe periods greater than 1 hour or 1 hour and a half. For practical reasons, it is not possible to continuously observe the satellite over a complete revolution. This would not only imply a considerable number of stations, observations over oceans as well as over continents, but also observations would have to be carried out under cloudy conditions, which is not possible at the present time.

In general terms, at infinity or at a very great distance, the earth's potential is the same as is produced by a point mass. The greater the distance to the center of the earth becomes, the more the problem is simplified. At the distance of the moon, only the first two or three harmonics have any effect.

UTILIZATION OF SATELLITES

Our knowledge of the earth's gravitational potential (Figure 2) was considerably increased over the last 10 years because of artificial satellites.

Since the satellite moves around the earth, its motion is influenced by variations in the gravity field at the altitude at which it is located. If we observe irregularities or perturbations in the motion, we can obtain important information on the gravimetric field or, which is the same thing, on the earth's gravitational potential. We can always represent a function by a sum of simpler functions. In the present case, we will represent the gravity field, which depends on the longitude, latitude and altitude, in the form of expansions into series of sine and cosine angle functions which depend on the latitude and longitude of the position at which one wants to calculate the value of the field. The coefficients depend not only on these two quantities, but also on the altitude. The determination of the gravity field therefore amounts to determining the value of the various coefficients which are called harmonic.

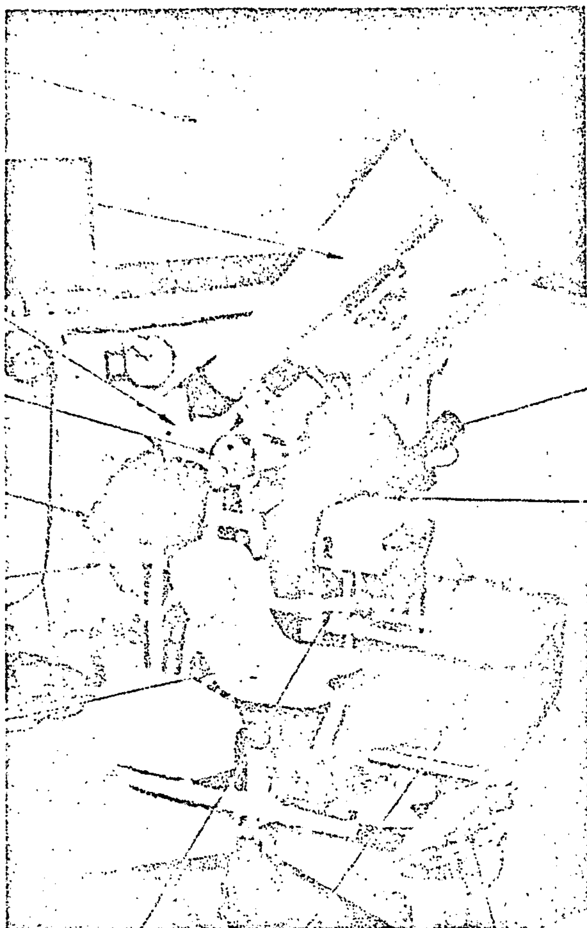
We can first construct a theory of motion of the satellite which allows us to describe the trajectory using the series mentioned above. Each harmonic of the gravitational potential will require several terms of the series. The theory can predict the period of these terms, but their amplitude remains unknown. Since the amplitudes are directly related to those of the coefficients of the gravitational potential, the analysis of the motion of the satellites will result in their values.

Sometimes an argument is used against this approach to the problem which says that this method in theory leads to the determination of an infinite number of coefficients. This of course is out of the question, and it is necessary to truncate the series. But this could have the result of falsifying the terms which are determined, unless there is a natural elimination, i.e. if it can be demonstrated that the eliminated terms are very small. This is generally the case.

On the other hand, the accuracy in the determination of the field is limited by the accuracy of the observations themselves. For example, if the data are accurate to 20 meters, we can only determine the orbital perturbations which have at least this magnitude.

Due to the results of this method, we can adopt the following point of view: if we have data with an accuracy of the order of 20 meters, we can determine between 80 and 100 harmonics of the gravity field. If the accuracy of the observations is improved by a factor of 4, i.e. if it is greater than 5 meters, we can determine 200 terms. If the accuracy reaches 20 cm, we can determine about 350 terms, etc. In addition to the accuracy with which the gravitational field can be determined, there is the resolution power, i.e. the distance below which we can detect details in the variations of the field because of the large altitude at which the satellites move. It can be shown that, using a solution with 100 coefficients, the resolution power is equal to 1800 km. With the 200 coefficient solution, it amounts to 1300 km, etc.

We have arrived at a stage where an increase in the accuracy of the observations is no longer significant for the gravity field. If we wish to improve our knowledge, other methods must be used. This also means that if we want to describe the motion of a satellite up to say, 20 cm, 350 parameters are sufficient for describing the influence of the earth's gravity field. The corresponding calculations are quite complex, but are easily carried out with large computers presently available.



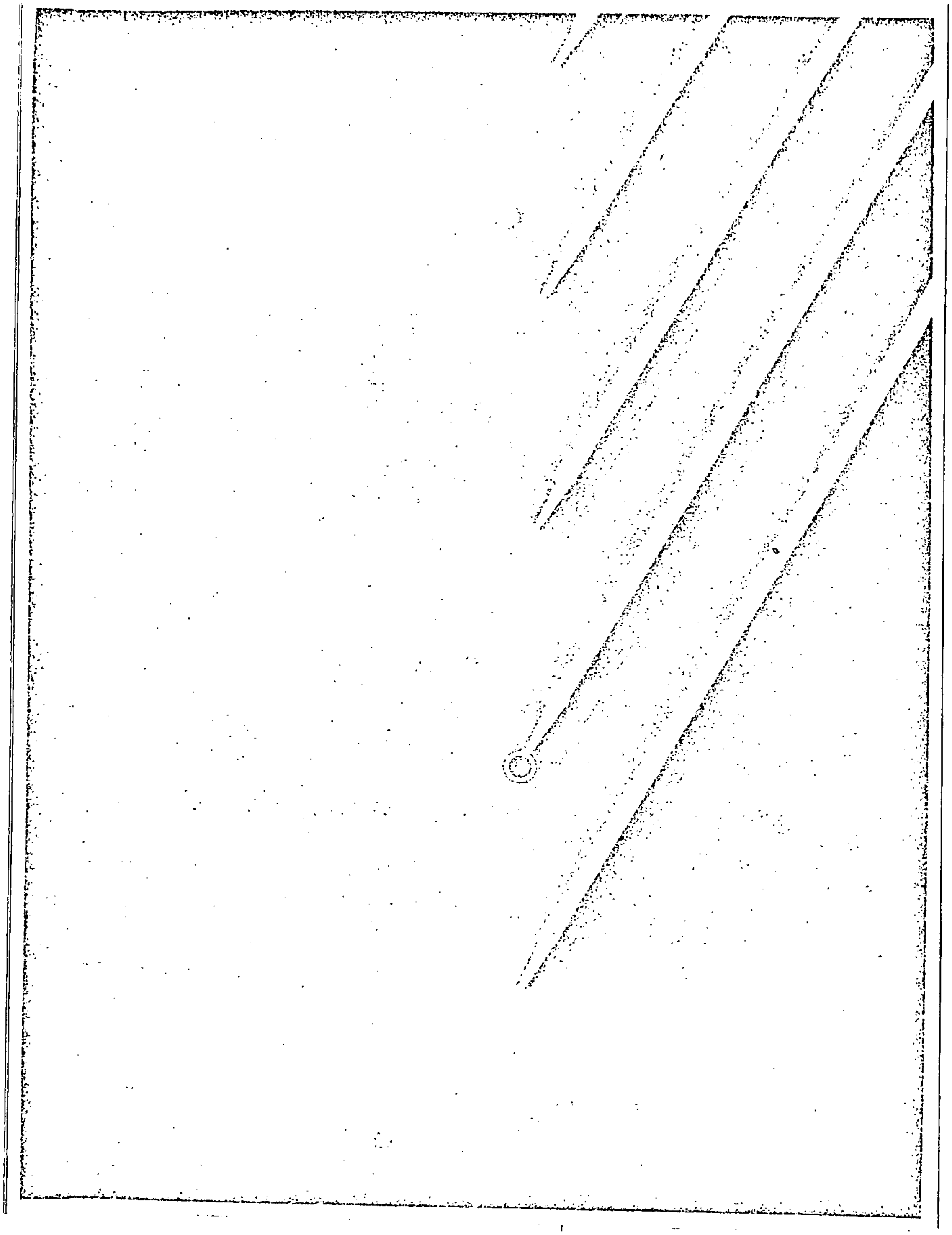
In order to know the position of a satellite, it is necessary to calculate its coordinates in the sky and its distance to the earth. If we photograph the satellite against a star background, its position with respect to them can be determined. The star coordinates are known from star catalogues. The accuracy of the direction of the satellite defined in this way varies between 1 and 2 seconds of arc, which is equivalent to a mean error of 10 to 20 meters in the position of the satellite along its trajectory. In order to calculate the distance of the satellite from the earth, a laser beam is transmitted in the direction of the satellite. The satellite reflects the beam. Then, the time required for the beam to complete the round trip is measured. The photograph shows a laser telemetry station built by the ONRS aeronomy service of the ONRS with the participation of the ONES and DRME.

AN OPTIMAL SOLUTION

The gravity measurements on the surface of the earth are the most direct method of determining the gravity field. However, this method requires considerable amount of time and a large number of observations over several tens of years, because of the large number of measurements to be made on the ocean and over the continents. In 1965, only 25% of the earth's surface was covered by observation stations. Considerable progress has been made since then, especially on the oceans, but a large global effort has yet to be made.

In contrast to measurements made from satellites, observations made at ground level are much more sensitive to variations in local gravity field than variations at a much larger scale. Also in order to determine the global field, we must first take the average of the local variations to then eliminate them in a certain sense. In order to be valid, this smoothing would require many observations over small areas. Only under these conditions can one define an average value which would make sense. The interest in gravity measurements /32
on the surface is due to the fact that they can give variations of the field within small areas, which our satellite measurements could never give. It therefore appears that a judicious combination of the two types of data will lead to an optimal solution.

The geophysicists E.M. Gaposchkin and K. Lambeck of the Smithsonian Institute Astrophysical Observatory gave a solution which is precisely based on such a combination. It includes more than 300 parameters, 170 of which are determined by the analysis of satellite trajectories and the rest are based on gravimetric data. In effect, the two types of data play an equivalent role /33
for many of the terms. Their analysis was based on more than 100,000 observations obtained with 25 satellites and which extend over a period of almost 10 years. They were taken from 30 observation stations. Among the satellites used, we should mention the two French Diadème satellites. Most of the data consist of photographic observations which correspond to accurate satellite positions, with an accuracy of 15 - 20 meters. Much of the laser data, having



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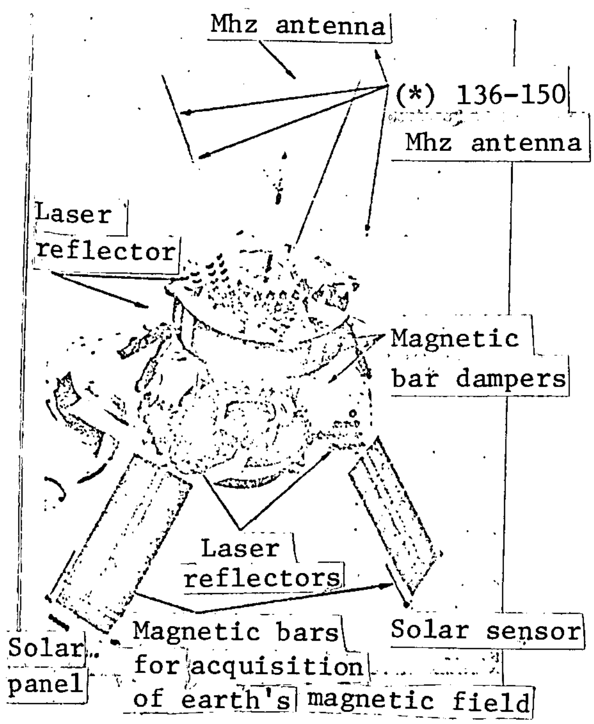

When several laser beams are directed towards the satellite, it is possible to not only calculate the distance to the earth but also its direction. The ONERA has developed a method which consists of the simultaneous use of two lasers: one transmits short pulses lasting about 30 nanoseconds (1 nanosecond - 10^{-9} second); the corresponding beam has a sufficiently small energy. It is used for telemetry. The other laser transmits pulses lasting on the order of 1 millionth of a second and the energy of the beam is much larger. It is used to calculate the direction of the satellite. The atmosphere disperses the light from these beams which leaves a trace which can be photographed (on this photograph, the trace of various beams transmitted by each one of these two lasers can be seen). This trace becomes weaker to the extent that the altitude of the beam is increased and the less effect the diffusion phenomenon has. There is a brilliant (surrounded by a circle) point which corresponds to the echo reflected by the satellite along the extension of the thickest beam. In order to avoid confusion of this point with a star, two pictures of the sky are taken which are slightly displaced in space. The second one is taken after the beam has passed. The stars then show up as two bright point in contrast to the laser echos which are represented by a single point (ONERA photograph).

an accuracy on the order of a few meters, was also used, particularly the data obtained from French stations.

Using data having an accuracy of 20 cm, we can obtain a resolution power on the order of 900 km, as we already mentioned, and a corresponding accuracy in the geoid shape (shape which the earth would have if all of the surface were covered with ocean) on the order of 10 or 20 cm. However, local phenomena have amplitudes which are much larger than the average accuracy of the global representation. Thus, the Puerto Rico depression is related to a height variation of the geoid amounting to approximately 10 m over a distance of 150 km. If it is desired to represent such phenomena by spherical harmonic expansions,

* Translators note: illegible in foreign text.

these expansions must include at least 30,000 terms. Considering the enormous difficulties in the solution of this problem, and it is not even certain that it can be solved, it is necessary to attack this problem in a completely different manner.

DETERMINATION OF THE SHAPE OF THE GEOID

One of the methods which seems very promising for determining the shape of the geoid with the best resolution is altimetry. Consider a radar covering a satellite. The satellite transmits a signal to the ground where it is reflected and comes back to the satellite. The delay between the transmission and the reception gives a direct measure of the distance between the satellite and the reflection point on the ground. This is simply the inverse of a classical radar station with which the motion of the satellite is followed.

Such measurements can be carried out over the oceans and continents, but it is the first case which is the most interesting, because not only do the oceans cover 75% of the surface of the globe and reflect signals which are more coherent and more accurate than those sent back by continents, but in addition these observations are directly related to the geoid. The shape of the earth and its gravity field describe this geoid.

If the satellite is attitude-controlled along the vertical of the point, selected in such a way that the transmitted radar signal is reflected by the point closest to this vertical, its altitude above the ocean can be directly measured. Thus, if the orbit of the satellite is known from observations of tracking stations on the ground, the altitude measurement allows the determination of the surface of the ocean, which corresponds to the geoid surface with an accuracy of almost a meter.

Using an altimetric satellite on a polar orbit, the height of the geoid can be obtained over 75% of the surface of the globe over a very short time compared with methods based on gravimetric measurements on the open sea, which are very long and have a poorer accuracy. The remaining 25% of the surface of the earth can be covered by gravimetric and astro-geodesic measurements in order to measure the details of the geoid. If the reference orbit and the altimetric data are accurate to about 1 meter or better, any detail with an amplitude of more than 1 meter can be measured on the geoid.

At the present, the calculations of the orbit are still not very accurate. However, the altimetric data having an accuracy of 1 meter can be utilized in conjunction with the observations from ground stations in order to improve at the same time the orbit of the satellite and the potential models. This is based on the same principles which are currently in use. If this approach were used to determine the earth's potential, and if an accuracy in the altimetric data on the order of 1 meter is assumed, we can utilize series developments which will include approximately 1,500 terms without the introduction of significant numerical problems. The altitude of the geoid would then be known to about 50 cm. With this accuracy, it would be possible to find the relationships between small tectonic phenomena and the gravity field and to confirm the correlations already found.

If this accuracy were improved further, one could begin the solution of numerous problems in oceanography, because the average level of the oceans is only the approximate shape of the geoid due to of the ocean currents, tides, waves, variations in atmospheric pressure and winds. Thus, at the same time and with the same degree of accuracy, we can observe the variations in the mean level of the oceans and study the same phenomena as oceanographers.

It thus seems that geodesy, celestial mechanics, and classical astronomy must be considered as techniques to be used in geophysics, geology and even oceanography. Thus, astronomers will not only come back to the earth, they will penetrate into it.

photography on a star background cannot be done if meteorological conditions are unfavorable, especially if the sky is cloudy. On the other hand, observations at radio frequencies can always be made. These observations allow one to determine the distance of the satellite with an accuracy of 5 to 10 meters, under the condition that there is a radar transponder on board. It is also possible to measure the radial velocity using the Doppler effect if there is a very stable frequency transmitter on board. This is the case for the American "Transit" satellites used for ship navigation. This is also true for the French "Diademe" satellites of which the satellite D1-D is shown here. The measurement accuracy of the radial velocity is on the order of a few centimeters per second (photo CNES).

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